

Extraterrestrial Lava Flows

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Introduction

Volcanism is a fundamental process shaping the surfaces of the terrestrial planets. During the last three decades diverse and often spectacular **volcanic** features have been revealed by spacecraft images. These features include not only lava flows and basaltic-type volcanoes, which are present on the **Moon, Mars, Venus, Jupiter's moon Io** and possibly on Mercury, but also structures interpreted as **pyroclastic** flows, **silicic** domes, products from "**ice volcanism**" and, in the case of **Io**, possible **sulphur** flows. Active extra-terrestrial volcanism has so far only been confirmed on **Io**, where it occurs on a vast scale, and on Neptune's moon **Triton**, where active geysers were observed by Voyager 2. Volcanoes on the Moon and Mars are thought to be extinct, in the case of Venus this question remains open.

One important difference between volcanism on Earth and on the other planets is plate tectonics. Because plate tectonics does not appear to have operated on other planets, with the possible exception of Venus, extra-terrestrial volcanic features can be quite different from those on Earth. For example, the much **larger** sizes of the volcanic edifices on Mars have been in part attributed to the lack of plate movement, allowing repeated eruptions on the same location to persist for a long time, building massive structures.

Most of the available data on extraterrestrial lava flows are morphological in nature, obtained by imaging systems aboard orbiting or flyby spacecraft. Limited data on surface composition have been obtained by spectral analysis using both ground-based instruments and experiments aboard spacecraft. However, the most desirable way of studying the compositions of planetary surfaces is by means of analysis of samples brought to Earth by sample collecting missions. The drawbacks of such missions, whether they are manned or unmanned, are their much greater complexity and considerably higher cost than the remote sensing missions. Because rock samples have so far only been collected and brought back from selected sites on the Moon, only a few in-situ measurements by lander craft have been performed on Venus and Mars, and spectral data on planetary surfaces in general have been obtained mainly at low resolutions, the study of extraterrestrial lava flows has relied largely on morphological data, using empirical and theoretical models which link conditions of emplacement to final morphology.

The following sections will review what is currently known about lava flows on the Moon, Mars, Venus and 10. Readers interested in a description of the types of remote sensing instruments deployed in spacecraft missions are referred to Greeley (1987). General reviews on the geology and volcanology of the terrestrial planets and satellites may be found in Taylor (1975, 1982), Guest & Greeley (1977), Johnson (1978), Carr (1980, 1981), Wilson & Head (1983), Carr et al. (1984), Greeley (1987), Saunders et al. (1991), and Head et al. (1991).

The Moon

Lava effusions have been extremely important processes on the Moon. The dark areas of the Moon's surface, called maria (Fig. 1), are results of extensive, flood-like basaltic eruptions which spanned a long period in the Moon's history, peaking between 3.9 and 3.2 billion years ago. Mare basalts are mainly found on the lunar nearside, where they flooded older impact basins such as Imbrium. Studies of individual flows, however, are made difficult by the fact that few flow fronts can be distinguished on the maria, and their source areas are rarely identifiable. Most vents are thought to have been fissures which have subsequently been covered over by lavas. The scarcity of visible flow fronts and margins are possibly due to the lavas being too fluid to preserve the fronts, or because outlines have been eroded by numerous small meteorite impacts. Some flows on Mare Imbrium present clear outlines (Fig. 2) and can be traced back for several hundred kilometres, but even in those cases it is very difficult to identify their source areas, as the flows tend to overlap one another. Detailed mapping of the Imbrium flows by Schaber (1973) suggests that their source area was a fissure some 20 km long and that the late-stage lavas have travelled several hundred kilometres from their sources over gentle gradients.

Thickness estimates of mare flows have been diverse (Spudis et al. 1986), but experimental work on the mare samples (Brett 1975, Grove & Walker 1977) suggests that they were derived from relatively thin cooling units, only tens of metres thick. Such results are in good agreement with photogrammetric measurements and other photogeological studies (e.g. Schaber 1973; Moore et al. 1978) which indicate that the thicknesses of young flows from Mare Imbrium are usually of the order of 10 metres. Significantly thicker flow units were thought to have formed early in the history of the maria (Greeley 1976, Guest & Greeley 1977) and these may have required hundreds or even thousands of years to cool and solidify (Basaltic Volcanism Study Project 1981).

The difficulties in mapping individual lunar flows from source to toe and in obtaining reliable measurements of flow dimensions limits the application of both empirical and theoretical models to infer eruption parameters. Nevertheless, several attempts have been made to model the rheology of lunar lavas. Hulme (1973, 1974) and

Hulme & Fielder (1977) used a Bingham plastic model to calculate yield strengths of lunar flows and, in some cases, also derived effusion rates from channel morphology. Yield strengths for Mare Imbrium lavas were found to be of the order of 400 Pa and flow rates to be of the order of $2-8 \times 10^4 \text{ m}^3 \text{ s}^{-1}$ (assuming a Bingham viscosity of 10 Pas). Impact-melt flows associated with impact craters which are non-volcanic in origin, having been produced by heat from the impact process, generally showed higher yield strengths, up to $2 \times 10^4 \text{ Pa}$. Three impact-melt flows showed well-defined channel morphology which was used to derive effusion rates. These effusion rates were found to be generally lower than those derived for Imbrium lavas, ranging from $260 \text{ m}^3 \text{ s}^{-1}$ to $1000 \text{ m}^3 \text{ s}^{-1}$.

Moore and Schaber (1975) also used a Bingham model, pointing out that the existence of a Mare Imbrium flow lobe whose thickness ranged from 7-20 m on a gentle gradient (average 0.13 degrees) implied that the flow might have stopped because of its yield strength. They approximated the critical flow depth required by Hulme's (1974) formula to the flow's observed thickness and estimated the flow's yield strength to be 100-200 Pa, slightly larger than values measured for Hawaiian lavas, but lower than the value of 400 Pa reported by Hulme (1974) for a nearby flow. Moore & Schaber also estimated the minimum times for the flow to cool by radiation from the liquidus to the solidus (13 to 40 hours). In a later work, Moore et al. (1978) calculated the yield strength of several lunar crater (impact melt) flows which are non-volcanic in origin, having been produced by heat from the impact process which formed the crater. They used, whenever possible, the three Bingham model-derived equations to calculate the yield strengths of these flows and found yield strengths of the order of $2 \times 10^4 \text{ N m}^{-2}$, consistent with the values reported in Hulme & Fielder (1977). However, the use of the different equations resulted in an average difference of about 1.7 for the values of yield strength. After applying the same method to terrestrial (and martian) lavas, Moore et al. (1978) concluded that both lunar and martian lavas are more akin to terrestrial basalts than to terrestrial andesites, trachytes, and rhyolites.

Moore et al. (1978) did point out that application of Bingham fluid concepts to flows measured by remote methods must be done with caution, since it is still not clear whether the Bingham fluid model is valid for lava flows. Even if the model is valid, measurement problems remain, for example, Hulme's basic assumption that the final channel morphology is representative of the flow's initial passage through it may be flawed, as channel morphology is subject to change by the variable dynamics of magma supply and flow field development, such as flow breaching (Wadge & Lopes 1991).

The high extrusion rates derived by studies such as that of Hulme and Fielder (1977) may imply a basic difference between terrestrial and lunar magmas and crustal environment. However, calculations by Wilson & Head (1981) comparing the ascent and eruption of basaltic

magma on the Earth and Moon show that the differences between terrestrial and lunar magma theologies and **crustal** environments do not lead to gross differences between the effusion rates expected on the two planetary bodies for similar-sized conduits or fissures. Therefore, Wilson & Head (1981) argued that the presence of very long lava flows on the Moon (and, by implication, possibly very high discharge rates) suggests only that tectonic and other forces associated with the onset of some lunar eruptions were such as to allow wide fissures or conduits to form. They calculated that the surface widths of elongate fissure vents need be no wider than 10 m to allow mass eruption rates up to 10 times larger than those proposed for terrestrial flood basalt eruptions (e.g. 3×10^4 kg S⁻¹ m⁻¹ proposed by Swanson et al. (1975) for the Columbia River basalts), and that 25 m widths for the fissure vents would allow rates 100 times larger. Fissures only 10 m wide would be difficult to locate in most of the currently available orbital images unless characteristic features such as constructional **pyroclastic** deposits or collapse depressions are present.

Sample Analysis

Laboratory analysis of samples collected and returned by the Apollo missions have vastly improved our understanding of lunar volcanism in terms of chemistry and mineralogy. Remote spectroscopic studies from the Earth (e.g. Pieters et al.. 1978) and from the Apollo X-ray and gamma-ray experiments (e.g. Davis & Spudis 1985) have also contributed to our understanding of the mineralogy of the lunar volcanic surfaces. The Moon's lack of an atmosphere enables measurements of the surface to be made directly, which aids the interpretation of remote measurements.

Lunar samples collected from the maria have made possible a variety of laboratory studies to determine chemical composition and magma rheology. However, such theological studies refer to magma viscosity rather than yield strength as, so far, no reliable method of estimating yield strength as a function of temperature and composition is available (Wilson & Head 1981). Estimates of viscosity from laboratory simulations of lunar lavas have been made (Murase & McBirney 1970a) and values of about 10 poise (1 Pa s) at 1400 °C were obtained. These low viscosity values, together with thermal conductivity studies of simulated lunar **basalts** (Murase & McBirney 1970b) which show that the heat loss from the active lunar lavas would have been very small, suggest that the lunar **basalts** may have been able to flow for long distances, even over the shallow slopes of the maria.

In terms of composition, analyses of mare lavas revealed that they are basaltic, but also that there are important differences between the lunar and terrestrial **basalts** (e.g. Taylor 1975), notably the absence of detectable H₂O in the lunar samples, and the higher abundance of iron, magnesium and titanium. This may imply a low silica content, perhaps 38 to 42% as was found for the Apollo 17

basalts (LSPET - Lunar Sample Preliminary Examination Team 1973). More peculiar are some samples collected from the Oceanus Procellarum region by Apollos 12 and 14, which are characterized by enrichments in incompatible elements, including potassium (K), the rare-earth elements (REE), and phosphorus (P). These nonmare basalts have been designated KREEP basalts, but only a few of the KREEP samples are thought to be igneous rocks crystallized from internally generated melts. Most available KREEP samples are impact-melt rocks and impact breccias, formed as a result of the extremely high temperatures and pressures involved in a crater-forming meteorite impact. However, there is growing evidence that KREEP contaminated many, and maybe most, lunar magmas as they oozed towards the surface (Binder 1982, Warren & Wasson 1980).

Differences in texture between the terrestrial and lunar lavas are also evident from sample analysis, which have shown that the lunar lavas lack the alterations found in terrestrial basalts due to chemical weathering and hydrothermal activity. Sample analysis also indicates that most lunar lavas were derived from deep within the mantle, at depths ranging from about 150 km to as much as 450 km, and that the majority of lavas were erupted between 3.1 and 3.9×10^9 years ago. However, other studies suggest that lavas were erupted on the Moon before 4×10^9 years ago (e.g. Ryder & Spudis 1980, Taylor et al. 1983) and also as recently as 1×10^9 years ago (Schultz & Spudis 1983).

There is still some doubt as to the composition of possible volatiles in lunar magmas. Sample analysis and calculations by Sato (1976, 1977) suggest that an important source of volatiles in mare lavas was a chemical reaction between carbon and iron oxides at pressures less than 170 bars, which produced metallic iron, CO, and CO₂. Thermodynamic calculations suggest that CO becomes dominant as the pressure decreases. Sample analysis by Housley (1978) deduced that between 250 and 750 ppm CO was typically produced in the mare basalts as they erupted.

Small Scale Features

The lunar surface lacks large-scale volcanic constructs, such as shield volcanoes. The absence of these features may be due to the common occurrence of high effusion rate flows on the Moon, as recognizable individual shields can only build up when the mean distance flowed by the lavas from one source area is substantially less than the mean spacing between sources (Head & Gifford 1980). The relatively few shield-like structures recognized on the Moon (e.g. Greeley 1976) are small (a few km in diameter) in comparison with the lava flows, and are not considered to have been the sources for significant quantities of lavas. The same applies to other small-scale features such as domes and cones (e.g. Smith 1973). Some elongate fissure-like structures with widths of many tens to hundreds of metres have been proposed as fissure vents for the mare lavas (Schultz 1976, Head 1976) but, according to the

calculations by Wilson & Head (1981) these features are far too wide to represent the true widths of the fissure vents, and are more likely to be the result of collapse around the vent after the eruption ceases.

Apart from lava flows, the most distinct features seen on the maria are sinuous rills and mare ridges. Sinuous rills (Fig. 3) consist of winding channels which may have a rimless pit at one end. They are found mainly around the outer edges of the maria and are interpreted as collapsed lava tubes or drained lava channels (e.g. Taylor 1975, Greeley 1971, Guest & Greeley 1977). Sinuous rills can be used to help map flow directions and general source areas. In terms of size, they are considerably larger than their terrestrial lava-tube counterparts, as an example, Hadley Rille on Imbrium is over 130 km long and 5 km wide in places. Head & Wilson (1981) proposed that the size difference between sinuous rills and terrestrial lava channels and tubes can be accounted for by the difference in gravity between the Earth and the Moon and by the higher discharge rates attributed to the lunar lavas.

Mare ridges (Fig. 2), also called wrinkle ridges, are prominent mare features which can be tens of kilometres long. The current consensus of opinion is that the ridges are compressional features (e.g. Muehlberger 1974) and that the period of major ridge production was synchronous with, or closely followed, the emplacement of major mare basalt sequences (Pieters et al. 1980). However, others (e.g. Strom 1971) have suggested that ridges were formed by lavas erupted along fractures, or by a combination of volcanic and tectonic processes (e.g. Guest & Greeley 1977). Recent identification and analysis of a number of terrestrial analogues (Plescia & Golombek 1986) suggest that wrinkle ridges result from anticlinal folding above thrust faults that break the surface.

Lunar volcanism is not exclusively confined to the maria, and there is evidence that some highland units are also volcanic in origin. A prime candidate is the Apennine Bench Formation, a light coloured plains unit between the Imbrium and Serenitatis basins. Orbital geochemical data indicate that this unit is composed of KREEP basalt, and interpretation of the geology suggests that the unit was emplaced by extrusive igneous processes (Spudis 1978, Hawke & Head 1978, Spudis & Hawke 1985). However, no lava flows have so far been recognized in this unit.

Future Data Acquisition

The Moon still offers many puzzles which may go unresolved until new missions provide the data needed. Although ground-based work should not be neglected, as telescopic reflectance spectra of the Moon and laboratory spectral studies of lunar samples can still significantly advance our knowledge of the surface mineralogy, investigators are urging NASA to fund return missions to the Moon. Amongst those proposed is the Lunar Observer (Nash 1991), an

unmanned polar orbiting mission which will seek to extend our global knowledge of the Moon, particularly in terms of **geochemical** and geophysical studies. However, it must be stressed that great progress on the nature and emplacement of the lunar **lavas** can still be made with present data using comparative studies of lunar and terrestrial **lavas**. In particular, improved models of **flood** basalt eruptions could lead to a greater understanding of how the mare **lavas** were emplaced.

Mars

Mars has a richer variety of volcanic **landforms** and distinctive lava flows than the Moon, including some of the most spectacular volcanic edifices and flow fields in the Solar System. In general terms, the problems associated with the interpretation of martian **lavas** are similar to those discussed for the Moon, such as the difficulty in obtaining reliable flow dimension measurements. In particular, flows can rarely be mapped back to their source areas, and thickness measurements are subject to large errors. No samples have yet been returned from Mars, though it has been proposed that the **shergottite**, **nakhlite**, and **chassignite** meteorites found in Antarctica (called the SNC meteorites) are martian in origin (e.g. Wood & Ashwal 1981). Topographic and spectroscopic coverage are more limited for Mars than for the Moon and studies using spectral reflectance are made difficult by the fine, iron-rich **aeolian** dust which blankets much of the martian surface (Bell et al. 1989; Christensen 1982). However, the imaging data obtained by the two Viking Orbiter spacecraft and its predecessor Mariner 9 are comprehensive and at the time of writing we look forward to additional coverage by Mars Observer. Some of the presently available Viking images have resolution as high as 10 m/pixel, though commonly the resolution is about 100 m/pixel. The two Viking landers also obtained valuable data, performing in-situ experiments on the northern hemisphere plains which included analysis of samples by **X-ray** fluorescence techniques. It is generally thought that no crystalline rocks were sampled and that all the two dozen or so samples analyzed by the Landers consisted of **partly** consolidated, weathered soils. Results (Table 1) showed that these samples appear to be derived from mafic to **ultramafic** source rocks and thus are grossly similar in composition to terrestrial and lunar **basalts** (e.g. Clark & Baird 1979). More recently, Burns (1988) proposed that the present-day martian regolith is similar to terrestrial **gossams**, which are iron-rich oxidized cappings over sulfide-bearing rocks. Burns (1988) suggested that the martian gossams may have been formed by reactions involving iron-rich **ultramafic** rocks similar to terrestrial komatiites.

Geological units

The global image coverage of Mars shows that there is a marked dichotomy between the northern and southern hemispheres (e.g. Carr et al. 1977, Carr 1980). The northern hemisphere is formed by

relatively young lava plains dotted with volcanic structures, some of which are spectacularly large. The southern hemisphere appears to be much older, as shown by the heavily cratered terrain, and the volcanic structures located there also appear more degraded than those in the northern hemisphere. The dichotomy is accentuated by a difference in elevation between the two hemispheres, on average, the southern plains are 1 to 3 km higher than the Mars datum. Most of the northern hemisphere stands at elevations below the datum, the main exception being the volcanic provinces of Tharsis and Elysium. The Tharsis region forms a bulge some 8000 km across, with a summit elevation 10 km above the datum; while the Elysium regions shows a much smaller but still significant bulge. It is clear that the Tharsis bulge has played a major role in the tectonic evolution of Mars but its origin is still uncertain. It may be linked to mantle convection associated with the separation of the core (Carr 1981) .

Most of Mars' volcanoes are concentrated on the northern hemisphere in the Tharsis and Elysium regions. The most conspicuous volcanic structures are giant shield volcanoes, of which Olympus Mons is the largest known volcano in the Solar System (Fig. 4), being over 25 km high and some 600 km in diameter. Other giant shields are Ascreus, Pavonis, and Arsia Montes which sit atop the Tharsis bulge aligned SW to NE. These are morphologically similar to Olympus Mons, having shallow flank slopes (4 to 6 degrees), complex calderas and numerous lava flows discernible on the summit region and lower flanks. One significant difference, however, is that younger lavas from surrounding plains have buried the lower flanks of the three Tharsis shields, while the lower flanks of Olympus Mons stop abruptly at a scarp several kilometres high. The origin of the Olympus Mons scarp, along with that of the enigmatic corrugated terrain surrounding the volcano (the aureole) , is still uncertain. It is possible that they both resulted from mass movement removing the volcano's outer flanks (e.g. Lopes et al. 1982) .

Also located on the northern hemisphere is Alba Patera (Fig. 5), a peculiar, shallow structure some 1600 km across and 6 km high, topped by two nested calderas surrounded by graben (Carr et al. 1977, Mouginis-Mark et al. 1988). "Patera" is a collective term for a variety of unusual 'saucer-shaped' features which often have a central caldera (Greeley & Spudis 1981) . Some of the clearest and best-defined lava flows on Mars are seen on images taken around Alba's summit caldera and on the lower flanks (Cattermole 1990, Lopes & Kilburn 1990, Pieri & Schneeberger 1991).

Other volcanic structures found on Mars are domes, also named tholii. These are relatively small volcanoes which may, in some cases, have experienced explosive activity (Mouginis-Mark et al. 1982a) . Pyroclastic activity also may have formed several breached cones which are morphologically similar to cinder cones on Earth (Plescia 1981, Tanaka & Davis 1988), though some of these appear to

have lava flows emanating from them (Mouginis-Mark et al. 1992) . In addition to all the above, Mars has thousands of **sub-kilometre** sized hills on the northern plains (e.g. Frey & Jarosewich 1982) which are themselves volcanic in origin.

The volcanic plains on Mars are not unlike the lunar maria, having features such as wrinkle ridges (e.g. Sharpton & Head 1988), sinuous rills (e.g. Schaber 1982), and overlapping flow lobes, which are mainly found around the periphery of shield volcanoes. Detailed geologic maps of the flow lobes found on the plains have been prepared by Scott & Tanaka (1986). Many of these flows seem to originate several hundreds of kilometres from the large volcanic constructs (Schaber et al. 1978, Mouginis-Mark et al. 1982b) suggesting that vents, fissures and feeder dykes are quite numerous within much of Tharsis (Mouginis-Mark et al. 1992) . Source areas for specific flows, however, have not been identified in the presently available images.

Numerous lava flows lobes of several different morphological types are seen on the flanks of the large shield volcanoes. The major flow types are well represented on Alba Patera and were **originally** described by Carr et al. (1977). More recently, Scheenberger & Pieri (1991) produced a detailed map of the flows in this region (Fig. 5). The major flow types are: (i) sheet or tabular flows (Fig. 6), which can be several hundred kilometres long, (ii) tube-channel flows or crested flows (Fig 7), which have positive vertical relief with an axial apex which coincides with a valley and/or alignment of pits, and (iii) tube-fed flows presenting wide "leveed" marginal structures (Fig. 6), possibly similar to terrestrial leveed flows. The source areas of some of these flows on Alba Patera can be inferred (for example, some can be traced to the edge of a caldera) and therefore it is possible to obtain reasonably reliable estimates of flow length, area and widths (Lopes & Kilburn 1990) . However, even in those cases where the flow's outline and lobate front can be clearly seen, reliable thickness measurements are a major problem. Measurements of ground slope are also hard to obtain accurately. Topographic maps are available for the whole planet, but the topographic resolution is seldom sufficient on the scale of the sizes of most lava flows to determine the flow's underlying ground slope.

Lava Flow Studies

Several studies have attempted to relate the morphology of martian lavas to their conditions of emplacement or chemical composition. As for lunar flows, yield stress models have been applied to martian flows by Hulme (1976), Carr et al. (1977), Moore et al. (1978) and Zimbelman (1985) . Hulme (1976) applied the same technique he used on the Moon (e.g. Hulme, 1974) to a lava flow on Olympus Mons imaged by Mariner 9, relating the flow's levee width and local slope to the yield stress and, in turn, the yield **stress** to the lava's composition. Hulme's results indicated that the

lava's yield stress was in the range of $3.9 \times 10^3 \text{ N m}^{-2}$ to $2.3 \times 10^4 \text{ N m}^{-2}$ and its effusion rate between 380 and $470 \text{ m}^3 \text{ s}^{-1}$. The silica content estimated suggested that the lava was more silicic than typical Hawaiian lavas.

Using the higher resolution data acquired by Viking, Carr et al. (1977) used Hulme's technique on four leveed flows on the flanks of Arsia Mons and compared the results with those obtained using the alternative model proposed by Moore & Schaber (1975). Yield strengths obtained by both methods were found to be in reasonable agreement with one another, in the range of 10^3 to 10^4 N m^{-2} . Effusion rates, assuming viscosities of 10 to 10^3 N m^{-2} , were found to be in the range 10^3 to $10^4 \text{ m}^3 \text{ s}^{-1}$. Moore et al. (1978) applied the Bingham model (using 3 distinct equations) to martian flows and found the average yield strength for 11 Arsia Mons lavas to be of the order of 10^3 N m^{-2} , and those for 3 Olympus Mons lavas to be higher (of the order of 10^4 N m^{-2}), consistent with the results of Hulme (1976) for a different Olympus Mons flow. However, Moore et al. (1978) disputed Hulme's (1974) proposal that yield strength is simply related to silica content and argued that yield strength is partly a function of topographic gradient. As discussed earlier, several authors (e.g. Moore et al. 1978, Crisp & Baloga 1990b) have pointed out the uncertainties involved in using the Bingham model and, in particular, the use of channel and levee morphology to infer lava properties. Nevertheless, the Bingham model has continued to be widely used in planetary volcanology (e.g. Zimbelman 1985, Cattermole 1987). Cattermole (1987) found yield strengths for 9 Alba Patera lavas to be between 1.9×10^3 and $2.8 \times 10^4 \text{ N m}^{-2}$, comparable with the ranges quoted above and with Zimbelman's (1985) results for Ascreus Mons lavas, which ranged from 1.2×10^4 to $3.8 \times 10^4 \text{ N m}^{-2}$. The validity of these and other results mentioned above, however, can easily be disputed.

A new approach using the Bingham model was taken by Wadge & Lopes (1991), who proposed that the widths of the distal lobes of lava flows are representative of the rheology of the lava, assuming that the lobes represent the arrest of free-flowing isothermal Bingham fluids on a slope. Lobe widths are a useful practical index because they are typically about an order of magnitude larger than lobe thicknesses and can be measured far more accurately on remote images. Moreover, lobes do not suffer from the changes in morphology that channels undergo during an eruption. Wadge and Lopes found a positive correlation between lobe width and silica content of the lava (Fig. 7) which is predictable from the isothermal Bingham model. This correlation was used to investigate 20 flows on the flanks of Olympus Mons. After lobe widths were measured and normalized to those that would be expected on Earth, they were found to be largely equivalent to those expected for terrestrial flows with andesitic/basaltic silica contents (Fig. 7).

A different method proposed for determining the rheological properties of martian flows is the surface structure model of Fink

& Fletcher (1978) and Fink (1980). This model relates the size and spacing of festoon-like ridges on flow surfaces to lava rheology, thickness of the flow's thermal boundary, and applied stresses. Festoon-like ridges are seen on a variety of terrestrial lavas, ranging in size from centimetres, such as the ropes on pahoehoe lavas, to tens of metres, such as the ridges on rhyolitic flows. Because ridges on martian flows are similar in size to those on terrestrial flows with a high silica content, some workers have compared them to rhyolitic, dacitic (Fink 1980), and trachytic (Zimbelman 1985) flows. In order to determine whether festoon ridges could be used to place constraints on the composition of flows, Theilig & Greeley (1986) used two Icelandic basaltic flows as analogues for martian flows displaying festoon ridges. The martian flows under investigation included those located west of Arsia Mons (Fig. 8). Theilig & Greeley found the viscosities determined for the Laki (Iceland) and the Arsia Mons flows to be comparable, mostly between 10^8 Pa s and 2×10^{10} Pa s. They pointed out that, even though these are high viscosity values for basaltic flows, they could be obtained by decreasing temperature, increasing solid content in the magma, or decreasing gas content, all of which are related. They concluded, therefore, that since basaltic magma may have a high viscosity under specific conditions and large festoon ridges occur on some basaltic flows, ridge height and spacing may not represent compositional variations. Based on the morphological similarities between the Icelandic flows and the martian flows under consideration, Theilig and Greeley concluded that the martian flows they examined were emplaced as large sheet flows from basaltic flood-style eruptions. They proposed that the festoon ridges represented folding of the surface crust in the last stages of emplacement when viscosities were high, either due to cooling or to high crystallinity lava being erupted under low temperatures.

More recently, radiation cooling models have become the favoured method of relating flow dimensions to eruption conditions. Pieri & Baloga (1986) proposed two models of radiative cooling, one assuming a thermally well-mixed flow and a second a model assuming the lava flow to be "unmixed", that is, made up of a thermally homogeneous core covered by an infinitely thin crust. Radiation from the crust was characterized by a constant "effective radiation temperature" which needs to be determined empirically. These two methods were used by Cattermole (1987) to derive the effusion rates of several Alba Patera lavas assuming a range of initial temperatures. However, this approach has been criticized on several important points such as the assumption of an infinitely thin crust (Crisp & Baloga 1990a,b). Crisp & Baloga (1990a) have proposed a more refined model using a finite crust thickness and assuming partial core exposure at the surface which is more consistent with field observations. Crisp & Baloga (1990b) used this approach to calculate effusion rates of a flow on Ascreus Mons previously mapped by Zimbelman (1985) to be in the range of 10^1 and 2×10^4 m³ s⁻¹. Crisp & Baloga's model can only estimate the effusion rate with

a minimum uncertainty of one order of magnitude due to the dependency of effusion rate on parameters which must be estimated from empirical studies of terrestrial lavas, such as the fractional area of the flow's surface where the core of the flow is being exposed by cracking and overturning.

An approach consistent with Crisp & Baloga's method has been developed by Kilburn & Lopes (1991) in the form of a flow field growth model which relates measurable parameters (maximum length, L_m ; maximum width, W_m ; average thickness, H ; and average angle of underlying slope, θ) to the duration of flow emplacement (T). The absence of gravity and viscosity terms in the model's equation (due to the presence of these factors in both the length and width terms which are ratioed) makes it particularly attractive for use on extra-terrestrial lavas. Assuming that the extra-terrestrial lavas in question are aa or blocky, and that similar conditions of flow field growth apply for the Earth and other planets (see Kilburn & Lopes 1991, for more details), it is possible to use the flow growth model to calculate their duration of emplacement. Lopes & Kilburn (1990) did so for 18 well-defined lavas on Alba Patera for which the vent locations could be inferred. Their results indicated typical average effusion rates to be of the order of 10^4 to 10^5 m³ S⁻¹ and durations which ranged from a few days for the single-type flow fields to over 200 days for the only multiple-type flow field. The high effusion rates could reflect a combination of larger source pressures, lower magma viscosities, and larger fissure dimensions than on Earth (Wilson & Head 1983). Lopes & Kilburn (1990) estimated fissure dimensions for 3 Alba flows and calculated the average effusion rates per unit length of fissure (averaged over the duration of the eruption) to be between 5 and 15 m² S⁻¹, which are comparable to values for basaltic eruptions on Earth. Therefore, they concluded that the high effusion rate values could mainly reflect the large sizes of fissures on Alba and thus could not be used directly to infer lava composition by comparing with terrestrial examples.

Future Data Acquisition

The use of the lava growth model requires reliable measurements of flow length, width, thickness, and underlying slope. At present it is difficult to find lavas on Mars for which all these measurements can be made with a satisfactory degree of accuracy (Lopes & Kilburn 1990). However, improved images and topographic coverage of the martian surface will soon be acquired by the Mars Observer spacecraft. Apart from the improved imaging system, which will return data with resolution as high as 1.4 m/pixel in selected areas (Komro & Hujber 1991), the spacecraft also carries the Mars Observer Laser Altimeter (MOLA, Garvin & Bufton 1990). MOLA will have vertical resolution as good as 1.5 m which will allow the direct determination of heights of features such as lava flows and of ground slopes. Future missions are expected to address the question of the chemical composition of martian rocks, performing

further in-situ experiments and, eventually, bringing samples back to Earth for analysis.

Venus

Venus is of particular interest to planetary geologists because it is the only planet in the Solar System of similar size and mass to the Earth and which, therefore, may possibly have had a similar geological history to the Earth's. The **Magellan** spacecraft has recently revealed in detail the remarkable range of **volcanic** and tectonic features which dominate the Venusian surface. Since these features are permanently obscured from view by the thick cloud cover which completely shrouds the planet, it was necessary to use radio waves to map the surface. **Magellan** used a synthetic aperture radar capable of mapping details as small as 120 m across, many times better resolution than previously obtained by the 'earlier Soviet Venera spacecraft or from ground-based radar studies from Arecibo. Since **Magellan** started mapping in September 1990 an extraordinary variety of volcanic features has been revealed including some very extensive lava flows. It is clear from the available images that Venus has undergone significant internal activity which produced features such as volcanic **calderas**, domes, folded mountain ranges, and extensive fault networks.

The overall density of impact craters on the surface of Venus indicates an average age for the surface of about 400 million years, young by planetary standards. It has been proposed that this young age is due to the resurfacing of large areas of the planet within the last 10 million years by relatively rare volcanic events which poured out large quantities of lava. In some areas **active** volcanism may have occurred as recently as the last few million years, as indicated by the total absence of impact craters on these areas. Although there have been suggestions that volcanism is still active on Venus (e.g. Robinson & Wood 1992), the evidence is still inconclusive.

Types of volcanic features

Over 80% of Venus' surface is composed of volcanic plains and edifices and the remainder, which is predominantly composed of tectonically deformed regions, is likely to be deformed volcanic deposits (Head et al. 1991). The young age of the surface is reflected in the pristine appearance of many of the **volcanic** features, which **Magellan** data shows in remarkably sharp detail. Venusian volcanic edifices have a wide range of sizes and shapes but the most common are small shields generally less than 200 m high with diameters mostly between 2 and 8 km (Head et al., 1991). Also present on **Venus** are flat-topped, table-like **features**, and dome-shaped and cone-shaped edifices, all interpreted as volcanic in origin.

The fact that few discrete lava flows have been mapped on the

shields, even though this may be due to insufficient resolution or radar contrast, has led investigators to doubt that the shields are a significant source for the extensive intershield lava plains, such as the dark intershield plains on Guinevere planitia (Fig. 9) , or indeed that shields and associated lavas have significantly contributed to resurfacing (e.g. Aubele & Slyuta 1990, Garvin & Williams 1990). More extensive lava flows are found in association with some of the larger (and less common) volcanic edifices, such as Sif Mons (Fig. 10) , a 300 km diameter, 1.7 km high structure. Extensive lavas were emplaced from vents in and near Sif Mons, including a radar-bright flow some 300 km long and 15 to 30 km wide (Fig. 11). A similar flow (Fig. 11) appears to have emerged from a vent below the summit of Sif Mons and traveled some 400 km until it reached a topographically low region which caused it to turn (Head et al. 1991) . Numerous other, smaller lava flows are associated with Sif Mons.

The question of whether pyroclastic volcanism has occurred on Venus is an important one since the high atmospheric pressure of Venus is expected to inhibit the exsolution of volatiles from ascending magmas that leads to pyroclastic activity (unless the volatile content of the magma exceeds 4% by weight: Head & Wilson 1986). The identification of pyroclastic volcanism on Venus could, therefore, indicate that volatile rich magmas have been erupted. Head et al. (1991) suggested one possible site in Guinevere Planitia where a 20 km-diameter radar-dark unit appears to mantle the surrounding plains and may be the result of a Plinian-style eruption. Wenrich & Greeley (1992) identified five other sites which have radar-dark inferred ash deposits mantling radar-bright plains, and proposed that pyroclastic volcanism has occurred on these sites.

The presence of relatively steep-sided domes that resemble dacitic and rhyolitic domes on Earth may suggest the existence of more evolved lava compositions on Venus. The only data so far available on the surface composition of Venus has been obtained by five Soviet Venera landers. Geochemical data from all but one landing site indicated that the surface is similar in major element composition to that of tholeiitic and alkali basalts (Surkov et al. 1984, 1987). However, data from one of the Venera 8 site (e.g. Nikolaeva 1990) are more consistent with an intermediate to silicic composition. Magellan images show the location of the Venera 8 landing site (within about 1 degree of uncertainty) to be near a pancake-like dome similar to andesitic, rhyolitic, or dacitic domes on Earth (Head et al. 1991) . Numerous Venusian domes, including those located SE of Alpha Regio (Fig. 12) , were examined by Head et al. (1991) and reported to be typically under 25 km in diameter, having heights of 100 to 600 m and volumes of 50 to 250 km³, thus being larger than most rhyolite and dacite domes on Earth. Head et al. (1991) suggested that the large volumes may be due to the higher surface temperature on Venus, which may permit magma to extrude more efficiently and to cool more slowly (and thus travel for greater distances) than similar magmas on Earth (see also Head

& Wilson 1986).

Other volcanic constructs which characterize the Venusian surface are ring-like features called coronae which have been interpreted as products of local plume-like mantle upwelling (Solomon & Head 1990). Coronae are circular to elongate structures with diameters in the range 200 to 1000 km, characterized by annuli of concentric ridges surrounding an elevated centre (e.g. Stofen & Saunders 1990; Head et al. 1991). Coronae are usually also characterized by interior flows, domes, and small (20-50 km across) edifices and exterior flows (Stofen & Saunders 1990). Some coronae show evidence that volcanism was linked to the uplift and radial fracturing and that volcanism continued in both the corona interior and exterior after the formation of the annulus (Head et al. 1990).

Among the spectacular volcanic features on Venus are large sinuous channels, some of which resemble lunar sinuous rills. The Venusian channels typically have constant widths which range from 0.5 to 1.5 km, lack associated lava flow lobes or deposits and, in some cases, terminate in low-lying areas forming large plains deposits. They can be very long: for example, the sinuous channel in SW Guinevere Planitia (Fig. 13) is over 1000 km long (Head et al. 1990). Venus also boasts the longest channel so far found in the Solar System, Hildr Fossa, which is 6800 km long. These channels have been interpreted as resulting from low viscosity lava (or lava emerging at high effusion rates) becoming turbulent and thermally eroding a channel into pre-existing plains (e.g. Hulme 1973, Carr 1974, Huppert et al 1987). It is possible that thermal erosion processes on Venus are helped by the high surface temperatures (Head & Wilson 1986).

Studies of Lava Flows

Venus has a great number of lava flows, ranging from those forming lava plains similar to the lunar maria and the martian northern plains to well-defined, thick lava flows. The great lengths and volumes of some of these flows have led to suggestions that they may be similar to terrestrial komatiites and flood basalts (e.g. Roberts et al. 1992). The most dramatic flow field, which has tentatively been named Mylitta Fluctus, covers an area of approximately 300,000 km² (Campbell et al. 1991, Head et al. 1991, Roberts et al. 1992). This massive flow field (Fig. 14) comprises flows several hundreds of kilometres long with widths ranging between 30 and 100 km in the medial and distal parts of the flow field (Roberts et al. 1992). Some flows show well developed channels and levees. The source area for most of the flows was shown by Magellan to be a caldera some 40 by 20 km, the location of which had been previously predicted by Senske et al. (1991a) on the basis of Arecibo data. Roberts et al. (1992) have mapped the stratigraphy of Mylitta Fluctus, identifying 6 major eruptive episodes and some individual flows within those episodes. They found that measurements of individual flow lengths and widths to be

often impaired by flow superposition and that thicknesses of these flows could only be roughly estimated by examination of the regional topography. Although Magellan carries an altimeter, it is difficult to use its measurements to estimate flow thicknesses because of the large size of the altimeter's footprint (20 km diameter in the Mylitta Fluctus region). Roberts et al.'s use of local topography to bracket flow thickness values yielded thicknesses of 30-90 m for the distal region of the flow field and a maximum estimate of 400 m for the thickness of the proximal region. Based on preliminary measurements, as well as general flow morphology and presence of channels possibly formed by thermal erosion, Roberts et al. concluded that the lobate flows within Mylitta are basaltic in general composition, possibly emplaced at very high effusion rates or as high-temperature, low viscosity basaltic komatiites.

Due to the great uncertainty in flow thicknesses, interpretation of eruption style and duration are subject to great uncertainties. The further uncertainty about whether the lavas are aa or pahoehoe in origin also poses problems for attempts at flow modelling since emplacement models are not necessarily applicable to both types (Kilburn & Lopes 1991). Pahoehoe textures have been suggested for many Venusian flows based on polarization ratio (Campbell & Campbell 1991a), Venera landing site morphology (Garvin et al. 1984) and on radar backscatter cross sections (Campbell & Campbell 1991b). It is possible, however, that the most radar bright flows interpreted as pahoehoe may in fact be aa, as moderately rough surfaces are thought to produce the brightest flow signatures (Campbell & Campbell 1991b). Better means of correlating radar signatures to lava textures would greatly aid in the future application of flow models to Venusian flows, however, such studies must also await until more reliable measurements of flow thicknesses and local topography become available.

Markedly different lava flows from those at Mylitta Fluctus are the unusually thick units investigated by Moore et al. (1992) which issue from a volcano in the plains between Artemis Chasma and Imdr Regio (Fig. 15). Moore et al. (1992) applied two techniques that rely on geometric image distortions to estimate flow thicknesses and obtained thicknesses ranging from 133 m to 723 m for broad lobes with widths ranging from about 5000 m to over 40,000 m. These large thicknesses, together with the presence of regularly spaced ridges with large separations and with the large widths of the lobes, are indicative of evolved lavas with large silica contents.

Calculations of yield strengths by Moore et al. (1992) based on a model by Orowan (1949) which does not require topographic gradients to be known produced values ranging from 2×10^4 to 4×10^5 Pa, comparable to those for terrestrial rhyolite, andesite, and basaltic andesites calculated in the same way. However, Moore et al. warned that compositional inferences should be viewed with caution because yield strengths of flow lobes calculated with

Orowan's (1949) method and compared with silica content data by Wadge & Lopes (1991) show such scatter that a correlation between yield strength and silica content is no longer present. Other rheological calculations by Moore et al. (1992) indicated Bingham viscosities to be quite large (ranging from 1×10^7 to 8×10^8 Pa s) and consistent with compositions more evolved than basalt. Although flows with these characteristics are rare on Venus, they indicate that silica-rich effusive volcanism (and possibly associated explosive volcanism) has played an important role on the evolution of the Venusian surface.

At the time of writing, the Magellan mission is still collecting data and results are being published at a fast pace. The acquisition of stereo images should be of particular interest to studies of lava flows, as such images will allow more accurate determination of flow dimensions. Studies using radar properties are promising, for example, radar brightness variations among flows have been suggested as indicators of age (Kryuchkov & Basilevsky 1989), with bright, rough flows representing the youngest volcanic activity, though such variations are unlikely to be the result of aging processes alone (Head et al. 1991). It is possible that a better understanding of the factors which influence radar brightness may allow relationships to be identified between flow brightness variations, flow texture, and variations in factors indicative of flow age.

Io

Io (Fig. 16), the innermost of Jupiter's Galilean satellites, is of particular interest to the study of extra-terrestrial volcanism because it is the only body outside the Earth known to have active large-scale volcanism. Io's volcanic activity was first detected on images returned by the Voyager 1 spacecraft in 1979, which showed a spectacular plume rising about 200 km above the satellite's surface (Morabito et al. 1979). Another Voyager instrument, the infra-red interferometer spectrometer (IRIS), revealed enhanced thermal emission from parts of Io's surface (Hanel et al. 1979). Active volcanism was confirmed when the most prominent thermal emission detected by IRIS (about 17° C, in contrast to the surrounding surface at -146° C) was shown to coincide with one of the plumes. Higher resolution camera images obtained later showed numerous lava flows and calderas (e.g. Carr et al. 1979), but no evidence that the lava flows were still active.

Io is a small body, similar in size (3640 km in diameter) and density (3.5 grams/cm^3) to our own moon, so it requires a heat source other than radioactive decay to drive its volcanic activity. An explanation for the activity was, in fact, put forward shortly before the Voyager encounters. Peale et al. (1979) proposed that tidal heating of Io generated by the gravitational interactions of Jupiter and the other Galilean satellites would cause melting of Io's interior and possibly lead to active volcanism.

Eighteen weeks after Voyager 1's dramatic confirmation of Peale et al.'s (1979) prediction, new observations of Io's activity were made by Voyager 1's companion spacecraft, Voyager 2. Intense activity was still taking place but significant changes were detected between the two flybys, including the **cessation** of the largest plume, Pele (Smith et al. 1979), and the change in the appearance of an area some $10 \times 10^4 \text{ km}^2$ of associated surface deposits. Another plume, Loki, was found to have increased in size by about 50%, reaching nearly 200 km above the surface. A total of nine eruptive centres were identified by the Voyagers of which Loki has remained the most active.

Since the Voyager fly-bys, Io's activity has been monitored from Earth (Johnson et al. 1984, Goguen et al. 1988) by means of infra-red astronomical observations. These often take advantage of occultations of Io by Jupiter and by the other satellites, when timing the disappearance of each spot can give its surface position to a precision up to 100 km in one dimension. These observations can also be used to determine a hot spot's temperature and even to detect new spots (e.g. Goguen et al. 1988). Results obtained so far have shown that the Loki region is a dominant and persistent source of Io's thermal emission (Johnson et al. 1984, Goguen and Sinton 1985) though recent (1990) measurements (J. Spencer, pers. comm.) found Loki to have become fainter than during the Voyagers' encounters. Pele still seems to be dormant (Goguen et al. 1992) and it is now thought that Io has both persistent hot spots (such as Loki) and "Pele-class" activity, which is short-lived.

Sulphur or Silicates?

The relative importance of silicates and sulphur in the composition of Io's upper crust and volcanic melts has been a much debated issue since the Voyager flybys. Arguments favouring a composition predominantly of sulphur have been based on spectral and temperature data and on the surface colours shown by Voyager. On the other hand a predominantly silicate composition is supported by the topography of some of Io's surface features and by other temperature measurements. Io's size and density, similar to the Earth's moon, are also indicative of a predominantly silicate composition.

The strongest evidence for the presence of at least a thin layer of sulphur on Io's surface comes from spectral observations (both by Voyager and from the Earth) and from laboratory studies (e.g. Fanale et al. 1979, Nelson et al. 1980, Nelson et al. 1987, Smythe et al. 1979). Voyager's IRIS experiment obtained a spectrum of sulphur dioxide gas over the erupting Loki volcano (Pearl et al. 1979) and ionized sulphur has been detected in the Io torus (Broadfoot et al. 1979), the trail of neutral and ionized particles that Io leaves behind in its orbit. Based on Voyager data, Smith et al. (1979b) and Kieffer (1982) interpreted the eruptive plumes as

S and SO₂ geysers. Spectral data still provides strong evidence for the presence of **sulphur** on Io, although Hapke (1989) has recently argued that the only unambiguously identified **sulphur** species on Io is SO₂, and that all the observed spectra can be modelled by a combination of SO₂ condensates and **basalt**.

Io's surface **colours**, which were first revealed in the Voyager images as predominantly red, yellow, and **orange**, were cited as evidence for the presence of **sulphur**. Sagan (1979) attributed these **colours** to different anhydrous mixtures of **sulphur** allotropes, plus SO₂ frost and sulphurous salts of **sodium** and **potassium**. Pieri et al. (1984) argued that the sequence of **colours** displayed by the flows around the Ra Patera complex (Fig 18) was consistent with Sagan's interpretation, as supposedly cooler materials were found to be further away from the vent. Interpretations based on **colour** have, however, been contested on several grounds. Firstly, it was pointed out that the exact **colours** of **sulphur** can be drastically altered by the presence of even small amounts of other materials and by the temperature of heating, rate of cooling, and age of the **sulphur** compounds (e.g. Gradie & Moses 1983). Secondly, laboratory studies of sulphur in a vacuum carried out by Nash (1987) showed that the rapid quenching of high-temperature (and thus highly coloured) **sulphur** flow required by Pieri et al.'s (1984) model is not possible and that solid **sulphur** in a vacuum cannot preserve its original post-solidification **colour**. In addition, more **précise** calibration of the Voyager data indicated that Io's **colours** are yellowish-green, rather than orange-red (e.g. Young 1984). Although the new calibration does not rule out the presence of **sulphur**, it casts further doubt on interpretations based on **colours**. A further complication is introduced by the fact that it is not known whether Io's surface **colours** are also those of the **crustal** materials underneath, or whether they merely represent a thin surface layer perhaps only a few millimetres thick.

Initial measurements by the IRIS instrument on the Voyagers indicated maximum temperatures of about 600 K, consistent with the melting temperatures of **sulphur**. However, since lava bodies typically develop a cool crust as well as glowing cracks (e.g. Crisp & Baloga 1990), the IRIS-derived temperatures do not rule out a silicate composition. Indeed, Carr (1986) developed a model of silicate volcanism on Io with a range of temperatures (matching the infrared observations) which could be attributed to different degrees of cooling of silicate materials.

Recent ground-based observations have tended to support a predominantly silicate composition. Johnson et al. (1988) reported a temperature measurement of about 900 K for a large eruptive event on Io and Goguen et al.'s (1992) measurements of a particular hot spot yielded temperatures of over 1150 K. These temperatures are significantly above the boiling point of **sulphur** in a vacuum (715 K) and are consistent with those of silicate magma. Johnson et al.

(1988) pointed out that although these measurements rule out pure molten S as the major constituent of magma in the eruptions observed, possibilities other than silicate volcanism remain, as there are other **sulphur** compounds with boiling points higher than 900 K such as sodium polysulfides (Lunine & Stevenson 1985).

Further evidence in favour of a predominantly silicate composition for Io's crust is provided by the presence of high mountains, steep **scarps**, and caldera walls. Clew & Carr (1980) argued that, due to the low thermal conductivity and low melting point of **sulphur**, ductile behaviour would occur at depths of at most a few hundred metres for all reasonable values of heat flow in a dominantly **sulphur** upper crust. The occurrence of ductile behaviour at such shallow depths would prevent the formation of observed topographical features on Io, such as **calderas** with depths greater than a few hundred metres and **scarps** higher than 1000 m, as these would not be self-supporting if they were composed primarily of **sulphur**. However, they added that a silicate crust with several percent (but not tens of percent) **sulphur** included could satisfy both the mechanical constraints and the observed presence of **sulphur** on Io. They also pointed out that injection of silicate melts into **sulphur** deposits could be the cause of some of the plumes observed, as proposed by Reynolds et al. (1980). Eruptions of silicate magma into near-surface **sulphur** could also result in remobilization of **sulphur** to produce flows. The implication of Clew and Carr's analysis for the flows is that, while the observed relief for the sheetlike flows is compatible with the critical heights for **sulphur**, the presence of **calderas** (which appear to be their source) makes a dominantly **sulphur** composition for Io's upper crust unlikely.

Most planetologists are now tending towards the conclusion that both silicate and **sulphur** volcanism are taking place on Io. A likely scenario is one proposed by Carr (1985) in which injections of hot silicates from beneath a silicate and **sulphur** crust can remobilize the **sulphur** causing flows and lakes within **calderas** which are maintained in a liquid state by the underlying hot silicates. This idea has been extended by Lunine & Stevenson (1985) to explain the temperatures and thermal fluxes in the Loki Patera region in terms of a convective **sulphur** lake heated by an underlying magma chamber. This model, which takes into account the physical and chemical processes in convective **sulphur** lakes, provides a way to relate the results from ground-based telescopic infra-red observations of Io's surface heat flow to changes in the temperature of a **sulphur** lake and to variations in the output of silicate magma chambers in the crust.

Surface features and Lava flows

Over 300 vent areas have now been identified on the Voyager images, most appearing as dark spots a few tens of km across. In some cases higher resolution images of these areas show nearly black volcanic

calderas which reflect less than 5% of the sunlight. Extensive lava flows have been mapped originating from several of, these dark volcanic centres. Other geological features identified on Io's surface include long, curvilinear cliffs and narrow, straight-walled valleys a few hundred metres deep, as well as mountains several kilometres high and regions of layered terrain with extensive plateaus and mesas. Io has, however, a remarkable absence of impact craters, suggesting very high resurfacing rates (Johnson et al. 1979).

Schaber (1980) identified three major types of flow materials on Io which are associated with different vent morphologies: pit crater flows, shield crater flows, and fissure flows. Pit crater flows are seen in the immediate vicinity of pit crater vents, generally extending from one side of the crater as massive coalescing flows. The flows generally present extreme colour and albedo variations and can be traced as far as 700 km from the individual vents. Shield crater flows are associated with shield constructs (such as Ra Patera, Fig. 20) which are concentrated in the equatorial region of Io. These flows are typically narrow and have sinuous paths, which may indicate that they flowed over significantly steeper slopes than the pit crater flows. Absolute values for the slopes, however, are not known. Individual shield flow lobes can be traced as far as 300 km and colour and albedo can vary along their lengths. The last and rarest type of flow is that associated with elongate fissure vents. Schaber (1980) identified only four occurrences of fissure flows and suggested that some of these flows might be significantly thicker than those from pit and shield craters. Fissure flows may, therefore, have had considerably higher viscosities or yield strengths than the other types.

The uncertainties in the composition of surface materials, together with the poor morphological and topographic data currently available on all three types of Io flows prevents the application to Io of the empirical and theoretical flow models discussed in earlier sections. If the flows on Io are primarily composed of sulphur an additional difficulty in modelling them is presented as relatively little is known about the morphology, emplacement mechanism, and physical and theological properties of sulphur flows. Terrestrial sulphur flows are rare and only one example, the Siretoko-Iosan sulphur flow in Japan, was observed while active (Wanatabe 1940). Among the few other examples described are those on Lastarria volcano in Chile (Naranjo 1985) and the 1950 flow on Mauna Loa which was studied by Greeley et al. (1984) as a possible analogue to the flows on Io. The Mauna Loa flow is thought to have been caused by the heat from basaltic flows of the 1950 eruption re-mobilizing secondary sulphur deposits which had accumulated on the flank of a basaltic cinder-and-spatter cone (Skinner 1970). The re-mobilization of sulphur by hot silicates may have occurred on Io, however, Io's flows are orders-of-magnitude larger in size than the Mauna Loa flow. Greeley et al. (1984) suggested that molten sulphur could have flowed long distances on Io as a result of (i)

relatively low viscosities in the melting range, (ii) sustained effusion resulting from continued heating of the source area, (iii) relatively low heat loss in the Ionian environment, even for thin sulphur flows (see also Fink et al. 1983), and (iv) formation of flow tubes which effectively extend the source vent to the flow front. As a result of their fluidity and low melting temperature sulphur flows on Io may form relatively thin veneers over other flows (possibly silicate in origin) and surface features.

The discovery of flows on Io which are possibly composed of sulphur has motivated research into terrestrial sulphur flows, including studies of industrial sulphur flows (Greeley et al. 1990) which are produced under well constrained conditions and provide a much needed alternative to field observations. Greeley et al.'s studies of the rheologic properties of industrial sulphur flows showed that the flows remained mobile at temperatures below those at which sulphur has been observed to solidify. They attributed this mobility to small quantities of impurities that can change the physical properties of molten sulphur and proposed that similar changes on Io may allow flows to be emplaced over larger areas than laboratory studies might predict. The observations of industrial sulphur flows also showed that crusts were maintained on the surface of many of the flows for much of their emplacement history and that they possessed some buoyancy, probably due to trapped gases. If this is true for Io, where the trapped gases could be SO₂ vapour, development of such "unpredictably stable" crusts (Greeley et al. 1990) might allow flows to become efficiently insulated and, consequently, travel further.

An interesting part of Greeley et al.'s (1990) rheological studies of sulphur flows consisted of applying Hulme's (1974) model to obtain viscosities and flow rates which were compared to measured flow rates and published viscosity values. Greeley et al.'s (1990) results suggested that the Bingham model is only applicable in the latter stages of a sulphur flow when both the viscosity and yield strength increase in magnitude and the flow rate is low. It is possible that turbulence in the early stages of emplacement would make Hulme's model invalid, but Greeley et al. (1990) also pointed out that since the cooling history strongly affects the behaviour of these industrial sulfur flows, the use of a Bingham model may not be appropriate for them. This conclusion has also been supported by studies of a naturally occurring sulphur flow, as Naranjo's (1985) application of Hulme's (1974) model to the Lastarria flows predicted unreasonably high effusion rates. Other models relating morphology to flow emplacement described in earlier sections have not yet been tested on sulphur flows but it is unlikely that they would be applicable without at least some modifications. Specific models relating sulphur flows' morphology to emplacement characteristics are clearly needed.

Future studies

The Galileo spacecraft is at present on its way to Jupiter and due to arrive in the Jovian system in December 1995. It is likely that Galileo will find that there have been significant changes in the morphology of the **calderas**, mountains, and flows, and that new features will have been formed since the Voyager observations. Galileo carries several instruments which will greatly further our knowledge of **Io's** surface and volcanism, including the Near-Infrared Mapping Spectrometer which will obtain high spectral resolution data at a spatial resolution as high as 50 m. Such data should play a key role in determining the role of **sulphur** versus silicates. Other instruments aboard Galileo include the Solid State Imaging system, which will obtain images of selected areas on **Io's surface** with a resolution as high as 20 m per picture element, and the Ultraviolet Spectrometer which will map the erupting plumes. Together these instruments will provide a comprehensive data set from which many of the present questions regarding **Io's** volcanic behaviour may be answered. Meanwhile, further studies of **sulphur** flows on Earth would greatly aid in the interpretation of **Io** flows, particularly if some morphological criteria could be established allowing the distinction to be made between silica and **sulphur** compositions.

Summary

Studies of extra-terrestrial **lavas** can be categorized as follows: (i) theoretical studies relating effusive volcanism to planetary parameters (e.g. gravity, atmospheric density), (ii) interpretation of **lavas** mapped on spacecraft images and application of theoretical models which relate measurable morphological parameters to composition, rheology, duration, and emplacement history, (iii) spectral studies using ground based or spacecraft acquired data, and (iv) metrological studies based on available samples or on lander spacecraft measurements. This review has concentrated on category (ii) studies because, firstly, the predominant type of data available on extra-terrestrial **lavas** are imaging data and, secondly, to stress that the theoretical and empirical studies of terrestrial **lavas** which are the core of this book are of key importance in the understanding of extra-terrestrial **lavas**. It is important, however, to take into account the limitations of such models, the inherent assumptions within them, and the **complex** combination of physical, chemical, and geological variables which can make results of theoretical and empirical studies uncertain. Nevertheless, the use of compatible models relating morphological parameters to different aspects of flow emplacement (e.g. eruption properties and magma composition) could greatly advance the understanding of a planet's effusive volcanism at least on a local scale. The further development and verification of models of flow behaviour using terrestrial data would represent a major step towards the proper interpretation of extra-terrestrial **lavas**.

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TABLE 1. Chemical Composition of Martian, Lunar, and Terrestrial Samples
(from Greeley and Spudis, 1981)

Oxide, wt. %	Sample*					
	1	2	3	4	5	6
SiO ₂	44.7	42.8	44.48	45.03	37.79	49.34
Al ₂ O ₃	5.7	n.a.	11.25	7.27	8.85	17.04
FeO	11.38	21.09	19.66	6.82
Fe ₂ O ₃	18.2	20.3	3.00		...	1.99
MgO	3.3	n.a.	17.32	16.45	8.44	7.19
CaO	5.6	5.0	9.54	8.01	10.74	11.72
K ₂ O	<0.3	<0.3	0.40	0.06	0.05	0.16
TiO ₂	0.9	1.0	n.a.	2.54	12.97	1.49
SO ₃	7.7	6.5
cl	0.7	0.6	
Sum	91.8	n.a.	98.37	100.45	98.50	95.75

Initials n.a. mean not available.

*Samples are the following: 1, Martian sample S1, Chryse Planitia [Toulmin et al., 1977]; 2, Martian sample U1, Utopia Planitia [Toulmin et al., 1977]; 3, model Martian lava, calculated composition [McGetchin and Smyth, 1978]; 4, lunar mare basalt, Apollo 12 olivine normative 12009 [Papike et al., 1976]; 5, lunar mare basalt, Apollo 17 high-Ti 70215 [Papike et al., 1976; and 6, terrestrial basalt, oceanic tholeiite [Engel et al., 1965].

Figure Captions:

Figure 1: The nearside of the Moon, showing the dark areas (maria) which represent extensive, flood-like basaltic effusions, and the lighter-coloured highlands. (Mount Wilson Observatory photograph).

Figure 2: Oblique view across Mare Imbrium taken by Apollo 15. Imbrium, one of the Moon's youngest maria. Note the lava flow running from the top right to the lower left of the image, and the mare ridges running from the top left to the lower right. (15-1555) .

Figure 3: Sinuous rills are common on the lunar maria. Shown here is Hadley Rille, which is about 300m deep and can be traced for tens of kilometres. The image is approximately 160 km across. (Apollo 15-0587).

Figure 4: The largest volcano on Mars, Olympus Mons, which is about 600 km across and 25 km high. (Viking 211-5360).

Figure 5: Map showing the distribution of lava flows on Alba Patera, Mars. Flows are predominantly oriented radially to the central caldera complex. (From Schneeberger and Pieri, 1991, courtesy D. Pieri) .

Figure 6: Various morphologically distinct types of lava flows on Alba Patera. Left: A tube-fed (leveed) flow is indicated by the arrow, other flows clearly seen in this image are of the tabular type. The image is about 100 km across. (From Viking rectified photomosaic MTM 45107). Right: Sheet (tabular) flows dominate this massive flow field, but a less common tube-channel (crested) flow can be seen at the lower left (marked C) The image is approximately 180 km across. (From Viking MTM 45117) .

Figure 7: Relationship between average distal lobe width (w) and silica content. Left: Plot of w (units are $\log_{10} \times m$) with 1 standard deviation error bars against silica content for terrestrial lava flows, with compositions ranging from basalt (hollow triangles) to andesite (filled triangles), dacite (hollow squares) and rhyolite (filled squares) . Right: Olympus Mons average distal lobe widths normalized to the Earth's gravitational field plotted on the same scale as graph on the left. (From Wadge & Lopes 1991) .

Figure 8: Flow lobes located west of the Martian volcano Arsia Mons. The distinctive arcuate festoon ridges, oriented normal to the inferred flow direction, are up to 50 km high and spaced 100 to 400 m apart . (From Theilig & Greeley 1986, courtesy of R. Greeley).

Figure 9: Magellan radar image showing a cluster of volcanic shields in Guinevere Planitia. The image is about 120 km across.

(Magellan P38304).

Figure 10: Magellan radar image showing Sif Mons. Image is about 450 km across. The two long, bright flows extending towards the top of the image are shown in greater detail in Figure 13. Image is about 55 km across. (Magellan P38054).

Figure 11. Magellan images of two flows from Sif Mons. Left: Flow unit embaying small shields in the regional dark plains. Right: flow unit being locally diverted into small graben in dark plains near centre of image. (Magellan P38171, P38055).

Figure 12: Magellan image of domes, each about 25 km in diameter, with maximum heights of 750 m, located SE of Alpha Regio on Venus. (Magellan P37125).

Figure 13: Magellan image showing sinuous channel in SW Guinevere Planitia. The radar-bright margins have been interpreted by Head et al. (1991) as levees. (Magellan P36706).

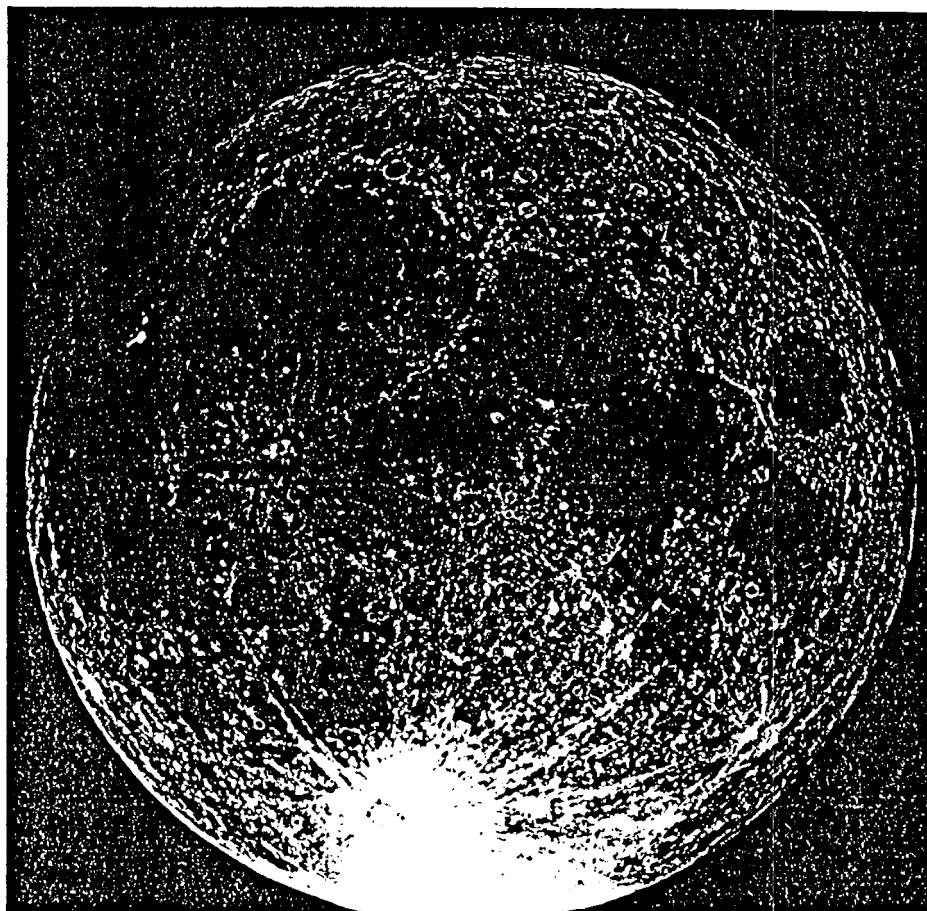
Figure 14: Magellan radar image of the massive flow field named Mylitta Fluctus. The image is approximately 900 km across. (Magellan P38289).

Figure 15: Magellan images of unusually thick lava flows on Venus from a volcano located on the plains between Artemis Chasma and Imdr Regio. White bar is 50 km long. Black strips and rectangles are missing image data. (Magellan MIDRP 37S164, courtesy of H. Moore) .

Figure 16: Full-disk image of Jupiter's moon Io, taken by Voyager 1. Io is the first extra-terrestrial body where active volcanism has been identified. (Voyager P-21457 C).

Figure 17: Voyager 1 image of Ra Patera on Io, showing the dark central caldera and surrounding flows. The total absence of impact craters suggests a young surface. The picture is about 1000 km across. (Voyager P-21277C).

Fig 1



100

Fig 2

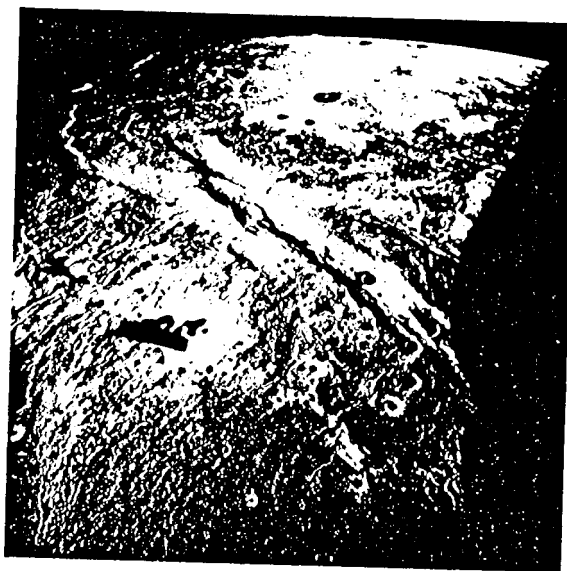
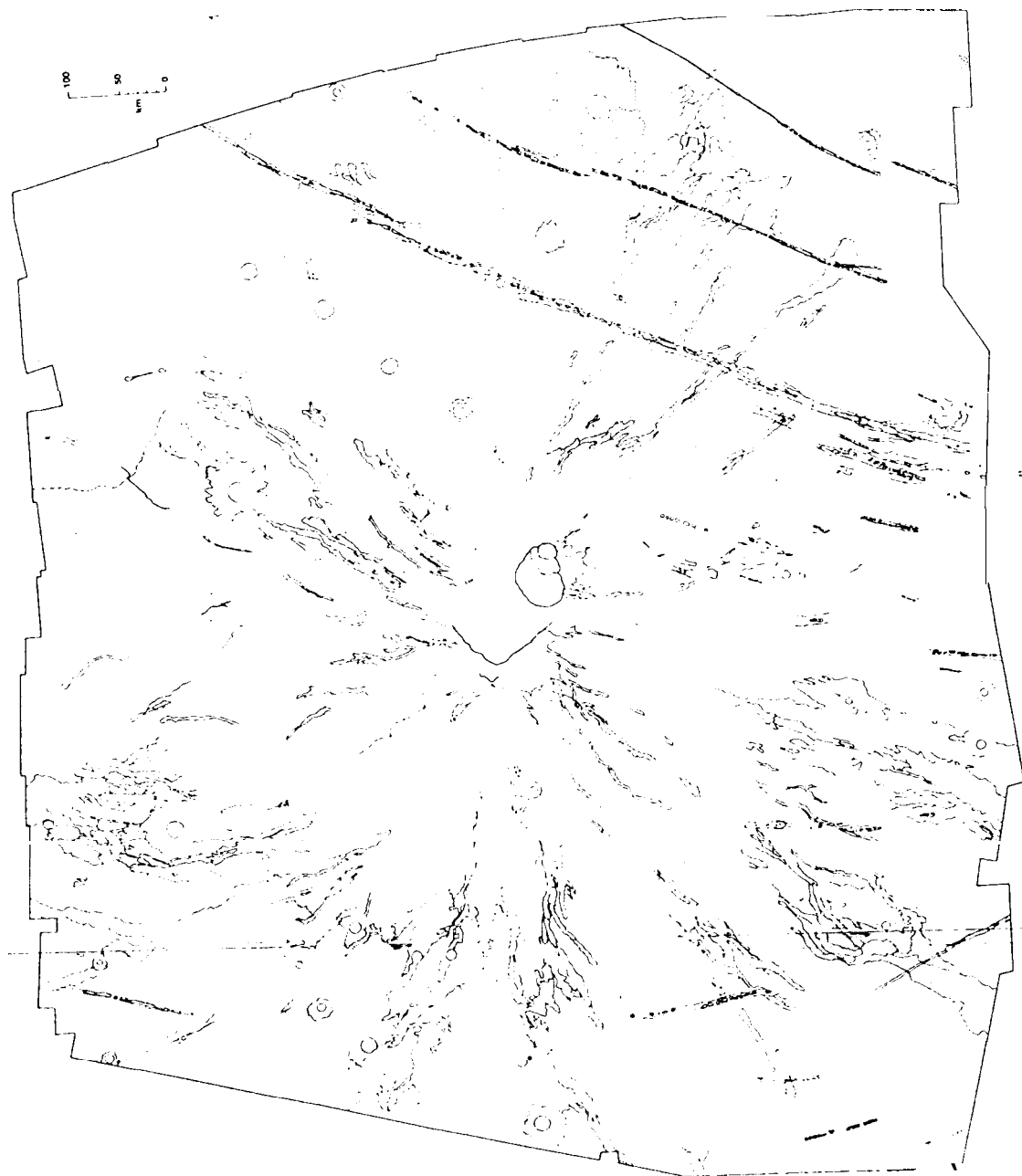


Fig 4



Fig 5



2 cm max

Fig 6

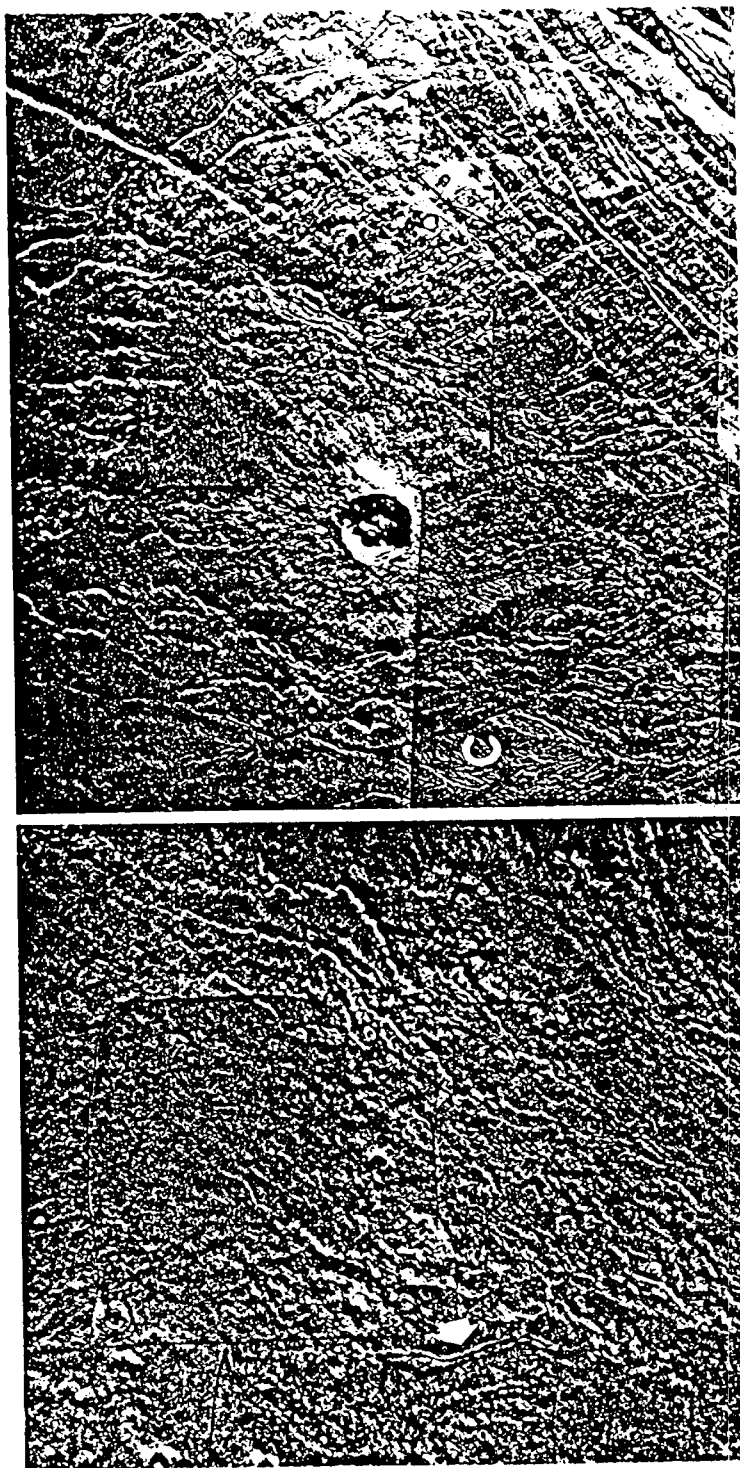


Fig 7

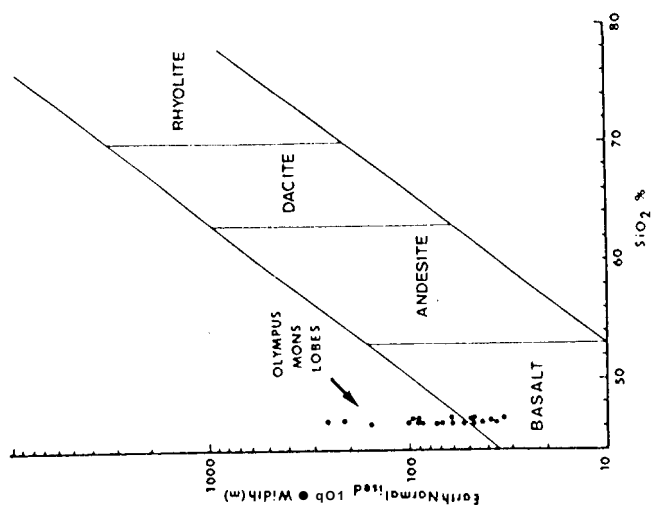
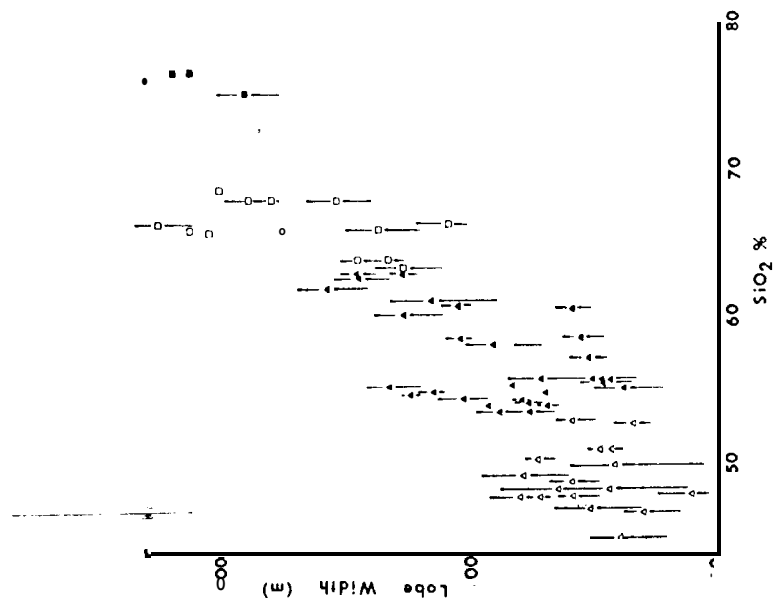
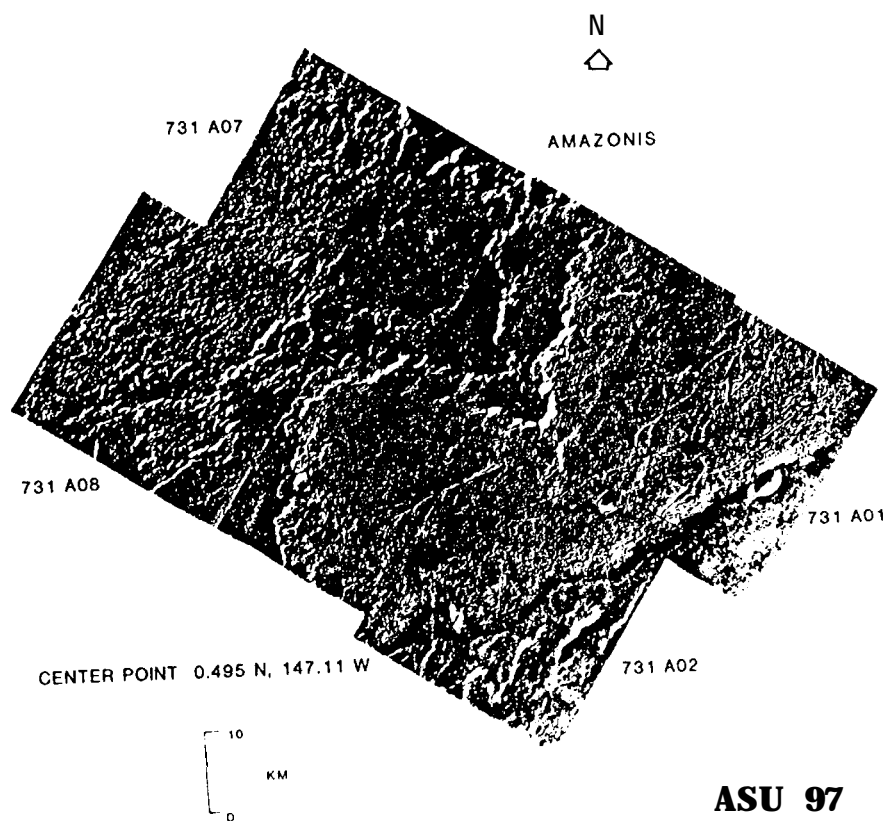


Fig 8



ASU 97

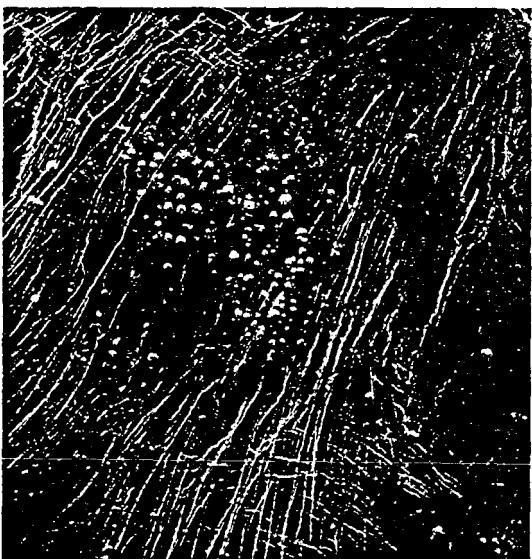


Fig 9



crop if
necessary

Fig 10

↓ crop
if necessary

Fig 11, left

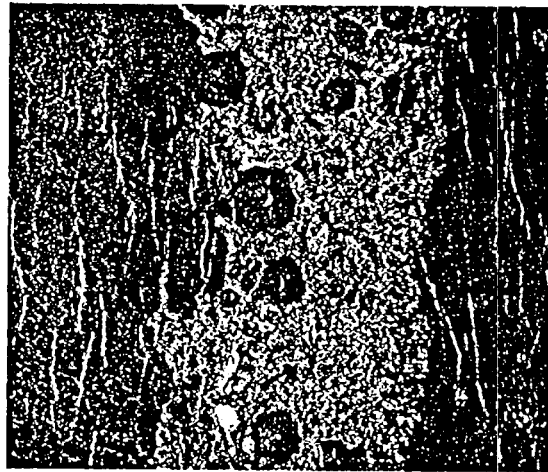


Fig 11
right hand side



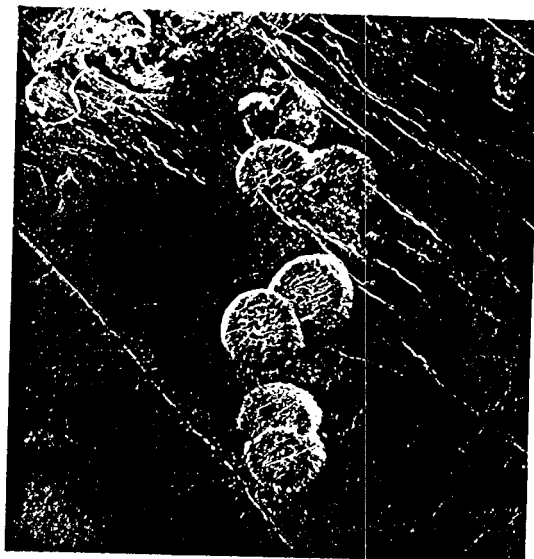


Fig 12

Fig 13

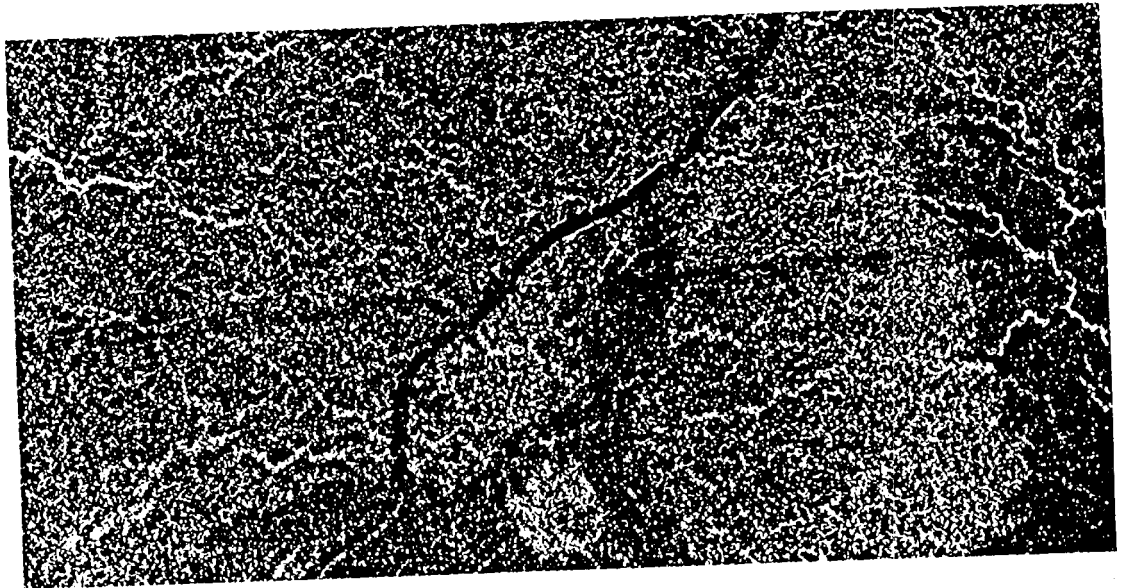


Fig 14



Fig 15

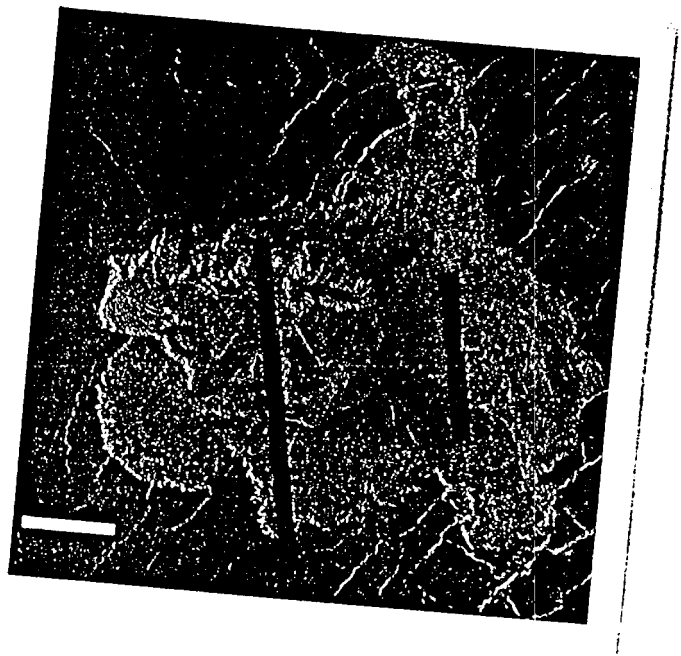


Fig 16

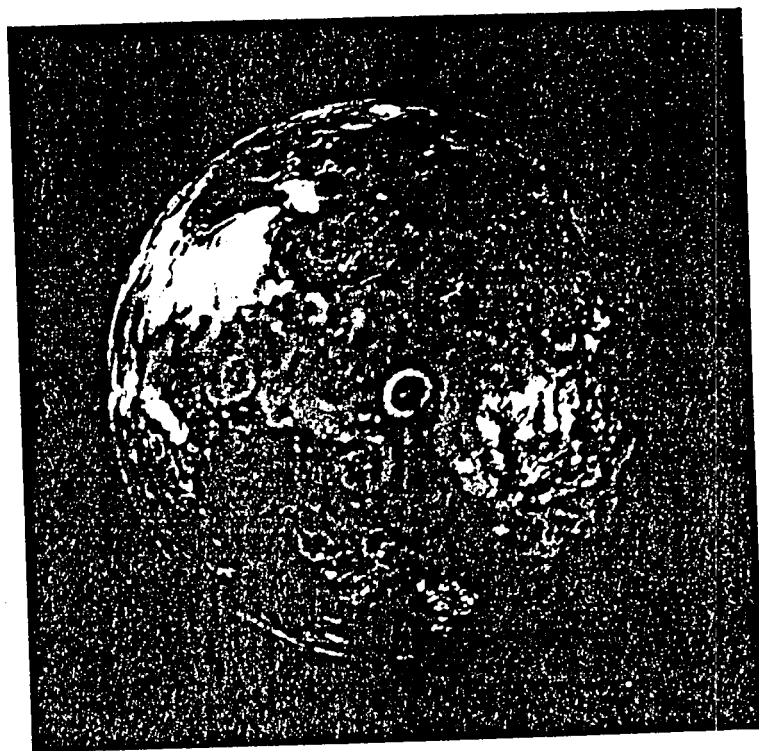


Fig 17

